THE ISOSTATIC STATE OF MEAD CRATER

W. B. Banerdt, A. S. Konopliv, N. J. Rappaport, W. L. Sjogren
Jet Propulsion Laboratory
California institute of Technology
Pasadena, CA 91109

R. E. Grimm
Department of Geology
Arizona State University
Tempe, AZ 85287-1404

P. G. Ford
Center for Space Research
Massachusetts institute of Technology
Cambridge, MA 02319

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Corresponding Author: W. Bruce Banerdt

Mail Stop 183-501
Jet Propulsion Laboratory
4800 Oak Grove Drive
Pasadena, CA 91109

Phone: (818) 354-5413
Fax: (8"18) 39?) -')22(,

bruce.banerdt@ccmail.jpl.nasa.gov

ABSTRACT

We have analyzed high-resolution Magellan Doppler tracking data over Mead crater, using both line-of-sight and spherical harmonic methods, and have found a negative gravity anomaly of about 4-5 mgal (at spacecraft altitude, 182 km). This is consistent with no isostatic compensation of the present topography; the uncertainty in the analysis allows perhaps as much as 30% compensation at shallow depths (~25 km). This is similar to observations of large craters on the Earth, which are not generally compensated, but contrasts with at least some lunar basins which are inferred to have large Moho uplifts and corresponding positive Bouguer anomalies. An uncompensated load of this size requires a lithosphere with an effective elastic lithosphere thickness greater than 30 km. In order for the crust-mantle boundary not to have participated in the deformation associated with the collapse of the transient cavity during the creation of the crater, the yield strength near the top of the mantle must have been significantly higher on the Earth and Venus than on the Moon at the time of basin formation. This might be due to increased strength against frictional sliding at the higher confining pressures within the larger planets. Alternatively, the thinner crusts of Earth and Venus compared to the Moon may result in higher creep strength of the upper mantle at shallower depths.

INTRODUCTION

Mead Crater, with a diameter of nearly 300 km, is the largest impact crater identified on Venus (Schaber *et al.* 1992). It presents perhaps the only opportunity for exploiting a crater to directly investigate the structure of the crust and lithosphere of Venus (and the effects of a large impact upon them) using gravity measurements, as the attenuation of short-wavelength gravity signals with altitude precludes the detection of signals from features significantly smaller than this. In November of 1992 Magellan's orbital periapsis passed near Mead Crater, making it possible to acquire high-quality, high-resolution Doppler tracking data at sufficiently low altitude to potentially resolve the gravity signature for an impact crater of this size. In this study we analyze the available gravity data over Mead crater and used it to constrain the isos atic state of this feature to better understand the properties of Venus's ithosphere.

Mead Crater

Although it is the largest impact structure yet identified on Venus, Mead was unknown prior to the acquisition of high-resolution radar images (Fig. 1a) and topography (Fig. 1b) by the Magellan spacecraft. It is centered at 57.2°E, 12.5°N, about 750 km northwest of the western "claw" of Aphrodite. It is situated in an area of rolling plains at an elevation close to the planetary mean. There is a grouping of coronae immediately to the west, and the plains surrounding the crater are heavily faulted. The deformation of the region is characterized by a diversity of styles, including distributed fracturing of the plains, concentrated deformation in linear belts, and radial patterns of faulting. Many of the latter appear to be connected with volcanic activity (domes, collapse structure, etc.).

Mead crater is classified as a multi-ringed basin, with an inner ring diameter of 195 km and an outer ring diameter of 280 km. Presumably, the inner ring represents the original excavation and the outer ring is a result of terrace slumping into the transient cavity (Schaber *et al.* 1992). Both rings are rather irregular, and appear to be composed primarily of inward-facing scarps (Alexopoulos and McKinnon 1992, Schaber *et al.* 1992). The rim-to-floor depth of the crater is about 1.8 km, with the floor within the inner ring being at nearly constant depth, sloping up to the rim relatively uniformly (at Magellan altimeter resolution) over the interval between the rings. The ejecta deposits are subdued, and the plains immediately surrounding Mead are relatively dark and featureless in Magellan radar images, indicating that they are mantled by smooth material. Whereas the local plains are heavily faulted, Mead itself does not show much evidence for tectonic disruption, with only faint lineations appearing on its otherwise flat and featureless floor. The interior of the crater appears brighter in radar illumination than the surrounding plains, and has an anomalously low emissivity (< 0.7), indicating either a different composition or strong volume scattering and low absorption near the surface at radar wavelengths (Pettengil *et al.* 1992).

Previous Gravity Investigations of Impact Craters

The gravity anomalies and isostatic states of large impact structures vary considerablyamong the terrestrial planets. For a Cl"at('J's topography (alone, with no contribution from subsurface density variations, a negative free- air anoma y (gravity at constant altitude) and zero Bouguer anomaly (gravity at constant altitude less the calculated contribution from the topography) will result. Isostatic compensation gives a smaller negative free air anomaly and a positive Bouguer anomaly. Attention was focused early on the lunar mascons (Mulle and Sjogren 19(,s). The positive free–air and Bouguer anomalies of larg (diameter $D \sim 400$-700 km) impact basins on the Moon have been well expla ned by a combination of high- density ma re basalt fill and mantle uplift (*e.g.*, Bratt *et al.* 1985a). In order to separate these two contributions, however, it was necessary to assume (without direct evidence) that the pre- mare topography was completely isostatically compensated by mantle uplift, and that the subsequent mare loading is completely supported by the flexural strength of the lithosphere. Nonetheless, the predicted thicknesses of mare fill are in reasonable agreement with estimates obtained by other methods, indicating that at least some mantle uplift is present beneath many lunar basins. Independent gravity modeling without such constraints was performed for the small lunar basin Grimaldi by Phillips and Dvorak (1981). Grimaldi (with inner and outer ring diameters of 200 and 400 km, respectively) is comparable in size to Mead. Uplift of the mantle beneath Grimaldi in this model is about 25 km and corresponds to 60–85% compensation of pre-mare topography.

In contrast to the large basins, smaller impact craters (D = 30- 130 km) on the Moon have zero or even negative Bouguer anomalies. These are primarily comparatively young craters, and the deviations from the expected positive Bouguer anomalies are likely due to the effects of uncompacted impact breccia (Dvorak and Phillips 1977).

The gravity signatures of small- and intermediate-size terrestrial impact craters are similar in character to their lunar counterparts. Small craters display negative Bouguer anomalies due to filling by uncompacted, low-density materials from ejecta fallback and transient cavity collapse, and somewhat larger structures show no Bouguer anomalies (Dvorak and Phillips 1977, Pilkington and Grieve 1992). In contrast to the large lunar basins (for which the models discussed above would predict substantial positive Bouguer anomalies even if they had not been subsequently filled with high-density mare basalts), two of the largest impact structures on Earth, Sudbury and Popigai, have Bouguer anomalies near zero (Popelar 1972, Manaytis *et al.* 1976). Several others have broad, ring-shaped negative anomalies with narrow positive anomalies at their centers (the latter of which would not be resolved at Magellan spacecraft altitude). The Chicxulub structure, which has negligible topographic relief, has a broad -20 to -30 mgal (1 mgal = 10^{-5} m/s^2) Bouguer anomaly that can be explained largely as the result of density contrasts between consolidated sedimentary target rocks and later basin-filling materials (Sharpton *et al.* 1993). Substantial mantle uplift may be ruled out because it would require an unreasonably large density contrast between the target and fill materials, although the central anomaly of about +15 to +20 mgal is consistent with ~5 km of uplifted deep (~15 km) basement rocks in the central zone (about 50 km in diameter). Note that although the principal observed ring (presumably the excavation boundary) has a diameter of 180 km, Sharpton *et al.* (1993) argue from gravity data for an outer ring at $D \approx 300$ km, making Chicxulub a near twin to Mead. Manicouagan ($D \approx 100$ km) and Vredefort ($D \approx 60$ km) also have this type of signature. Manicouagan has a 40 km wide +10 mgal peak at the center of a broad -10 mgal anomaly (Sweeney 1978), and Vredefort's - 20 mgal low has a narrow +25 mgal Bouguer anomaly superimposed upon it (Slawson 1976). Both of these central gravity highs have been interpreted in terms of a few km of uplift at mid- crustal (~10 km) depths.

The results for Martian impact structures are ambiguous, mainly due to the uncertainties in the thickness of the crust and in the gravity models themselves. Sjogren and Wimberly (1981) derived an apparent depth of isostatic compensation (ADC) of ~130 km for the Hellas basin. A comparable value (~115 km) was inferred for the Antoniadi crater (Sjogren and Ritke 1982). If the ADC is interpreted in terms of simple Airy compensation, then the Martian basins are 100% compensated in a 115-130 km thick crust. However, the crustal thickness of Mars may be considerably less than this. Global analysis of gravity and topography (Bills and Ferrari 1978) places a lower bound on mean crustal thickness of about 30 km (limited by requiring the thinnest inferred crust on the planet, in the Hellas basin, to have a thickness greater than zero). Line- of- sight gravity studies of Olympus Mons and Elysium Mons provide a range of permissible thicknesses ranging from 30–150 km, depending on the degree of isostatic compensation assumed (Janle and Ropers 1983, Janle and Jannsen 1986). A range of 30–150 km is also consistent with the results of the flexural loading study of a number of features by Solomon and Head (1990). If the ADC is much larger than the actual crustal thickness, then an increasing proportion of topography must supported regionally (flexurally) instead of isostatically. Thus the large ADC's for Hellas and Antoniadi would also be consistent with small Moho uplifts beneath these basins if the mean thickness of the Mart an crust is considerably less than 100 km. The Isidis Basin (D = 240 km) has a distinct positive Bouguer anomaly, similar to lunar mascon basins. I has also been inferred (by analogy) to have a similar structure to the lunar basins, with an isostatic mantle uplift and a flexurally supported volcanic load (Sjogren 1979).

From the preceding summary it appears that considerable ambiguity still exists in our understanding of the isostatic compensation states of impact basins and large craters. The best evidence for a significant amount of compensation (though still equivocal) is for lunar basins. Partial compensation may be present in some Martian basins, whereas mantle uplifts beneath terrestrial structures may be ruled out. After placing Venus into context with the other planets through our gravity analysis of Mead, we will discuss the mechanisms responsible for the presence or absence of isostatic compensation in the structure of large impact craters.

GRAVITY DATA

The gravity data used in this investigation is derived from a single source: the Doppler velocity tracking of spacecraft in orbit around Venus. The reduced form in which it is interpreted, however, comes in two flavors: a spherical harmonic (SH) representation of the global gravity field, and profiles of line-of-sight (LOS) Doppler acceleration residuals (with respect to a reference field) along selected spacecraft orbit tracks. The SH model represents the best gravity field fit to the aggregate tracking data set and provides a two-dimensional view of the field, but does not reflect all the information available in the raw data. The LOS profiles retain nearly the full resolution of the tracking data, but are subject to individual orbit errors and a variable "viewing" geometry that is more difficult to interpret than a more conventional vertical gravity field.

In this study we have used both representations of the gravity field. This was done primarily to gain the greatest confidence possible in the interpretation of the gravity signal for Mead. In addition, we were interested in how well the SH techniques, which have previously been applied only to the largest-scale features on the planets, would work when applied to a relatively small impact basin utilizing the newly-acquired high-resolution harmonic models.

Spherical Harmonic Model

The SH model used in this study, MGNP60FSAAP, is a 60th degree and order field (3717 coefficients) that incorporates Doppler tracking data from both the Pioneer Venus and Magellan missions (Konopliv and Sjogren 1994). On a global scale, the individual gravity coefficients are generally well-determ ned to about degree 30. However, th aggregated coefficients beyond degree 30 give a igh- delity represe t on f e local gravity in areas over whicl the Magellan spacecraf altitude was below about km, with a half-waveleng a spa ial esolution of about 300 km Additionally, the resol ion near periapse (including the vicinity of Mead crater) should be optimum, as ere is no a priori constraint on e g avi y signature. The gravity field is determined by ap lyi g only a s a ia a pri ri constraint for ose local areas where e surface gravity is not wel determ ed, minimizing he distortion of the field n r gi ons wher the data are str ng (see Kon p v Sjogren 1994).

The question that arises naturally is whether a 60^{th} degree field, containing no information at wavelengths shorter than 600 km, can adequately resolve a crater less than 300 km across. Clearly, not all of the gravity signal from the crater will be represented. But a significant fraction of the signal power is contained in the longer wavelengths. igure 2 shows the computed vertical gravity at the surface to 60^{th} degree and order from a simple topographic model of Mead crater, consisting of an cylindrical hole that has a depth of 800 m to a radius of 90 km, decreasing to zero at a radius of 140 km. The peak computed gravity at the surface is 28 mgal. Although this is only a third of the 90 mgal free-air anomaly that would be measured by a gravimeter on the surface, it is still a significant signal that contains useful nformation about the feature. At an altitude of 182 km the discrepancy is even less, w nearly half of the computed 13 mgal anomaly contained in the 60^{th} degree field

Figures 3a and 3b show the vertical acceleration at spacecraft altitude (182 km) and the surface, respectively, over a 15° square centered on Mead, corresponding to the same area shown in Fig. 1. Several features of these maps are notable. The gravity highs in the west and southeast are associated with a field of coronae and Aphrodite Terra, respectively. The gravity trough that runs south to the center of the figure, then turns to the east is associated with fracture belts. Of particular interest to this study is the low located near the center of the figure at 14°N, 59.5°E. This low straddles the northeastern edge of Mead and the edge of an adjoining low-lying region to the northeast. The offset of this feature from the center of Mead is within the resolution of the gravity model, which is well constrained in this region, and appears to be a combined effect of these two topographic features. Note that the anomaly in the surface field is more closely aligned with the crater, suggesting that the contribution from he crater is stronger in the higher frequencies than that from the feature to the northeast.

The amplitude of the anomaly with respect to its surroundings is difficult to determine, as there is a strong regional gradient with distinct contributions to the field from several adjoining features. In the field at altitude it appears to be at least -2 mgal, ınd perhaps as great as -4 mgal, whereas the anomaly at the surface appears o be 12 to - 20 mgal deep.

Line-of-Sight Residuals

The primary gravity data set consists of 13 X-band Doppler radio tracks, spanning 21 orbits from 6178 to 6198 on November 12 and 13, 1992. Five orbits had no tracking data and three orbits at S-band frequency having much higher noise levels were not included. Ten had a definite signature for the crater, whereas the three orbits at the beginning and end of the span had essentially no crater signature (6178, 6197, and 6198). The orbits crossed the crater in a north-south direction (85.5° inclination to the Venusian equator) and the spacecraft was at an altitude of 182 km over the crater. The LOS angle (the angle between the spacecraft-Earth line and local vertical) was about 24° in a WSW direction. The quality of the X-band data was excellent, having an average noise of less than 0.2 mm/sec for 2 second data samples.

The data were reduced in several ways to assure a valid determination for the crater anomaly. Reductions were made with different Venusian gravity field models: GM only (corresponding to a spherical reference), 21 degree and order spherical harmonics (McNamee *et al.* 1993), 36 degree and order harmonics (Nerem *et al.* 1992), and 60 degree and order harmonics (Konopliv and Sjogren 1994). Figure 4 displays LOS acceleration signatures obtained from the Doppler residuals for six of the orbits based on a reduction using the 21 degree and order model. The negative anomaly at 13°N corresponds very nearly to the center of Mead. Relative to the shoulder at 16°N, we infer that the crater anomaly at spacecraft altitude is approximately -5.0 ± 0.5 mgal for orbits 6185-6187. If the shoulders at 9°N and 16°N are assumed to be part of the background, the peak anomaly is about 3.5 mgal. Figure 5 displays the results for orbit 6186 when GM only and a 36 degree and order field are used. All three results (GM, 21, and 36) are very similar. This is due to the fact that the crater anomaly is a short-wavelength feature and these models have relatively low spatial resolution and are not able to resolve it. However, when a 60 degree and order reference model is used, a quite different result is obtained. An anomaly of at least -2 to -4 mgal can now be seen in the harmonic gravity model, although it is offset from the crater's center (see above). But the maximum amplitude of the LOS anomaly at 12.5°N after fitting the orbits with this harmonic model is only about –1 mgal, as shown in Fig. 6. This demonstrates that the 60th degree SH model has assimilated nearly all the signal available from the Doppler data in this region, and thus should be able to accurately (at the 1 mgal level) represent the gravity signal inherent in the data.

The temptation is to take the simple sum of the anomalies in Figs. 3 and 6 (although this would require a number of rather shaky assumptions). However, a two-dimensional interpretation using a contour map of the complete set of LOS residuals over this area relative to the 60th degree model (Fig. 7) shows that this is not warranted. It can be seen that the strongest remaining signal in the acceleration data (± 1.1 mgal) is associated with Mead. This residual anomaly is in the form of a dipole, with positive and negative peaks of equal magnitude closely adjoining each other. This indicates qualitatively that the total power in the SH anomaly is very nearly that contained in the Doppler data, as the closely-spaced positive and negative peaks will tend to cancel each other's effect over the broad anomaly. It also suggests (again qualitatively, from simple superposition) that the negative anomaly is located slightly too far (about a degree) to the northeast in the SH model, relative to the Doppler data.

Thus the anomaly associated with Mead appears to have a magnitude of between –3.5 to –5.5 mgal in the LOS data, and between –2 and –4 mgal in the harmonic model. However, it is difficult to assign a reliable amplitude to the Mead anomaly in either case, as it is superimposed on other strong signals in the region. In the following analysis we will attempt to better separate the crater signature from the regional background, and will compare the resulting anomaly magnitudes with simple crater models.

ANALYSIS OF GRAVITY DATA

Most previous studies of planetary gravity were required to use LOS analyses for all but the broadest features, due to the severely limited resolution of the harmonic models (*e.g.*, 18th degree for Venus [Bills *et al.* 1987], 18th degree for Mars [Balmino *et al.* 1982], and 16th degree for the Moon [Bills and Ferrari 1980]). However recent advances in computational power, analytical techniques, and tracking data quality have combined to make much higher degree and order fields possible (*e.g.*, McNamee *et al.* 1993, Nerem *et al.* 1993, Smith *et al.* 1993, Konopliv *et al.* 1993). The resolution of these models approaches the size of individual features of geophysical interest, opening the possibility of high-resolution gravity analyses using the harmonic models alone. In this investigation we have attempted to use both types of analysis in order to get the maximum value from the Doppler data.

Spherical Harmonic Analysis

In order to interpret the gravity over a small feature such a Mead, high-resolution topography data is required as well. For this study, we have used the most complete set of altimetry data available, which combines Venera, Pioneer and Magellan (Global Topography Data Record GTDR.3;1) data (Rappaport and Plaut 1994). The inclusion of cycle 3 data has resulted in excellent, nearly uniform coverage over this area.

The topography computed from the 360th degree model is shown in Fig. 8. This synthetic topography model reproduces very well the observed topography. The maximum depth of the crater is 870 m below the planetary mean radius (6051.84 km) and the floor of the crater is almost flat inside a circle of diameter 170 km. In the immediate vicinity of the crater, topographic lows of comparable depths are present to the northeast, northwest and southeast of the crater and topographic highs are present to the north, south, and southwest. It can be seen that although Mead crater is of considerable geological and geophysical interest, it does not stand out particularly from the surrounding topography, making it a difficult object to study.

In order to separate the Mead anomaly from the background signal, we applied techniques in both the spatial and wavenumber domains. We first did a regional SH analysis of Mead topography alone. The harmonic topography coefficients to degree 360 were determined for a planet modeled as a smooth sphere of radius R_t with the observed topography from the region $10° \le \phi \le 15°$, $55° \le \lambda \le 60°$ (translated vertically to a zero mean) superimposed, where λ and ϕ are latitude and longitude, respectively. We then computed the gravitational accelerations at spacecraft altitude associated with this local model. When the gravity is computed at a 60th degree resolution comparable to the observed field (Fig. 9), the Mead anomaly blends completely with that from the adjoining low, and the combined anomaly has a magnitude of -3 mgal, comparable in both position and magnitude with the feature seen in the global gravity model (Fig. 3a).

From this analysis we conclude that the assumption of uncompensated topography for Mead crater and the surrounding small scale structures leads to an acceleration above Mead crater nearly as large, relative to the surroundings, as the observed acceleration anomaly. Hence, Mead crater appears to be largely uncompensated.

In order to isolate Mead's signature using spectral techniques, we first computed the generalized isostatic anomaly map for the region (McNamee *et* *al.* 1993). This field (which contains that part of the gravity field that is uncorrelated with topography) reflects local departures from a globally averaged (by wavelength) apparent degree of compensation. Figure 10 shows the isostatic anomaly map for the Mead region calculated using the gravity and topography through degree 60. It can be seen that there is a prominent negative anomaly centered on Mead, indicating that an appreciable gravity signature corresponding to the crater is present in the 60^{th} degree model. Thus Mead is considerably less compensated than the global average at comparable wavelengths. (Note that Ovda Regio, to the south, also has a large negative isostatic anomaly, whereas it appears to be more, rather than less, compensated than other highlands regions [Herrick *et al.* 1989, Grimm 1994]. In this case the isostatic anomaly is negative because Ovda is a positive topographic feature rather than a negative feature like Mead.)

In order to more quantitatively assess the degree of compensation, we used the insight gained from the generalized isostatic anomaly, along with information obtained from the global admittance, to generate a map of the isostatic anomaly at constant depth. The spectral admittance of the gravity and topography is shown in Fig. 11. It can be seen that the global ADC is a strong function of wavelength, making it difficult to use a constant-depth isostatic gravity map to separate a feature from the background. However, at harmonic degrees higher than 30, the ADC is relatively constant, varying only between about 50 and 25 km, with the shorter wavelengths clustering around 25 km. In order to eliminate the long period background, we first applied a crude high-pass filter to the fields by retaining only the harmonics with degrees greater than or equal to 30. This removes all signals with wavelengths greater than 12° of arc (about 1250 km). Recall that from the LOS analysis (*e.g.*, Fig. 5), we can expect virtually all of the crater's signal to be contained in the harmonics with degree greater than 36. We then computed the deviation from isostatic gravity at the surface assuming a compensation depth of 25 km (Fig. 12).

A distinctive anomaly of -27 mgal can be seen centered on Mead, with the amplitudes of other features in the area considerably reduced. This means that the amplitude of the measured gravity signal is 27 mgal greater over Mead than expected if all local topography were compensated at a shallow depth of 25 km. The close correspondence of this value to that computed above for an uncompensated crater (see Fig. 2) further strengthens our conclusion that Mead is nearly uncompensated

Local LOS Modeling

A simple approach to the isostatic analysis is through direct comparison of observed local LOS gravity anomalies and those predicted from forward modeling of local topography and any compensating mass at depth. It was shown above that Mead is associated with a free-air anomaly of roughly -4.5 mgal at spacecraft altitude and a width of perhaps six or seven degrees of latitude along the spacecraft trajectory. We isolated the local topography of Mead by applying a linear ramp between distances of 1.3 to 2.0 crater radii. In other words, topography within about 200 km of the crater center was unchanged, that greater than 300 km was set to zero, and that in between was linearly attenuated.

Compensating mass anomalies at this scale are of course likely to be due to variations in crustal thickness and so an independent estimate of the depth to the Venusian mantle beneath Mead must be provided. Previous estimates range from 10-30 km, based on geodynamic modeling studies (Zuber 1987, Banerdt and Golombek 1988, Grimm and Solomon 1988). However, recent suggestions of a higher creep strength for diabase under anhydrous conditions (Mackwell and Kohlstedt, 1993) may call for an upward revision of these estimates. The global admittance estimates shown earlier (Fig. 11) suggests that wavelengths less than 1000 km are supported at a near-uniform depth of about 25 km. Grimm (1994) found that several plateau highlands (features specifically thought to be Airy-compensated) are supported on a crust that is regionally 30-40 km thick. Here we assume a reference crustal thickness $H = 20$ km. Because the spacecraft altitude (~180 km) is large compared to the estimated range in crustal thickness, results are insensitive to this parameter: a 10 km difference in crustal thickness from the reference value results in a change in LOS gravity anomaly of only about 0.1 mgal. The compensating masses at the crust-mantle boundary are taken to be opposite in sign to those of the surface relief but with magnitude f_c of the topography, where f_c is the compensation fraction.

The orbit simulation program GASP (Grimm 1991), patterned after ORBSIM (Phillips *et al.* 1978), was used to determine the dynamic LOS gravity anomalies along the spacecraft orbit due to the crater topography and subsurface mass anomalies. All masses are represented as a series of discrete points spaced according to the gridded topography.

The predicted LOS gravity anomaly for the case $f_c = 0$ is shown in Fig. 13. The peak-to-peak magnitude is ~4.5 mgal, and the shoulder-to-shoulder width is about 5 degrees of latitude. For 30% compensation at this depth the anomaly decreases to about -3.7 mgal. Assuming that these are projections into the LOS direction of purely vertical acceleration vectors, these values correspond to vertical anomalies of -5.3 and -4.3 mgal, respectively. Therefore the gravity data described above are well fit to a model of Mead crater that is at most 30% isostatically compensated.

DISCUSSION

Our evident hypothesis for the large negative gravity anomaly of Mead crater is that mantle uplift and isostatic compensation are largely absent here as beneath large terrestrial craters, notably Mead's terrestrial counterpart Chicxulub. In contrast, lunar basins (especially Grimaldi, Mead's best lunar analog) appear largely compensated.

The first problem this poses is how the (negative) load of the crater has been maintained to the present. If it is not isostatically supported, it must be supported regionally by the mechanical lithosphere, which allows some bounds to be put on lithospheric strength. With the help of reliable flow laws, one can derive thermal gradients and lithospheric thicknesses using viscous relaxation modeling (*e.g.*, Grimm and Solomon 1988). However, we can more simply put a lower bound on the mechanical lithosphere thickness by ignoring the history of modification and just looking at the present configuration, calculating the effective elastic thickness required to support the load today (note that Grimm and Solomon [1988] also considered flexural support in a manner very similar to what is presented here).

Using the thin shell formulation of Solomon and Head (1979), we calculated the vertical displacement w induced by a disk 1 km thick and 270 km in diameter with a density of -2.8 Mg/m^3, on a spherical lithosphere with Young's modulus of 1×10^{11} Nt/m^2 and values of lithosphere thickness varying from 10 to 100 km (Fig. 14). The maximum value $f_c = 0.3$ requires that the mass anomaly at the crust-mantle boundary, $\Delta m_m = w \Delta\rho$, be less than the compensation fraction times the mass deficit due to the surface topography, $\Delta m_c = t\rho_c$, where $\Delta\rho$ is the crust-mantle density contrast, ρ_c is the density of the crust, and t is the present depth of the crater, after the elastic response. This bounds the allowable displacement to crater depth ratio to be $w/t \lesssim f_c \, \rho_c / \Delta\rho$. Using values of $\Delta\rho$ = 0.5 and ρ = 2.8 Mg/m^3, we derive a limiting displacement ratio of 1.7, corresponding to a minimum lithosphere thickness, from Fig. 14, of about 30 km. Given that f_c may be much smaller than 0.3, Young's modulus is probably less than that assumed, and some viscous relaxation is almost certain, the thickness of the lithosphere is likely to be substantially greater than this.

We next turn to processes involved in the formation of the crater. Mantle uplift (if present) would most likely occur during the modification stage of crater evolution. Although this stage formally includes all morphological changes that happen after excavation of the transient cavity is nearly complete, we are primarily interested in distinguishing between short-term and long-term modification processes, and establishing in which one mantle uplift occurs. Changes that occur promptly after excavation determine the differences between simple and complex craters: simple debris sliding shapes the former, whereas large-scale collapse, forming central peaks, flat floors, and rim-wall terraces, are responsible for the latter. Basin rings (probably forming outside the original excavation) are also produced on this short time scale. Long-term alterations include slow floor uplift (viscous relaxation) as well as volcanic, tectonic, and erosional modification. Neglecting regional geological phenomena, impact crater modification is seen to be due mostly to the dominant driving force of gravity.

Melosh (1989, p. 160) assumed that mantle uplifts are formed by long-term viscous flow. However, the surface must also be displaced as isostatic equilibrium is established. This will cause doming of the crater floor and associated fracturing. These should be superposed on the "typical" basin morphology produced immediately after transient cavity collapse. Such a correspondence between inferred mantle uplift and basin topography and structure is not observed, although this constraint is problematic on the Moon, as subsequent volcanic burial may have obscured these features in the mare basins. The general lack of strong evidence for viscous rebound in large basins on the terrestrial planets and the Moon argues that either the viscosity has always been very high, preventing the establishment of isostatic equilibrium, or that equilibrium was established very early, obviating the need for subsequent relaxation. The latter hypothesis is supported by the apparent inverse correlation of mantle uplift with basin age and by viscous relaxation modeling (Bratt *et al.* 1985a, Solomon *et al.* 1982, Brown and Grimm 1993).

The alternative mechanism for mantle uplift is prompt displacement during transient cavity collapse. In this scenario the final basin topography is in or near isostatic equilibrium, which is the assumption made by Bratt *et al.* (1985a) in modeling lunar gravity and topography. Melosh (1989) concluded that a Bingham plasticoviscous rheology best describes crater collapse. Cohesion of a few MPa is inferred from lunar complex crater depths and terrace widths, whereas hydrodynamic rebound to form lunar central peaks requires effective viscosities less than $\sim 10^{11}$ Pa-s. Both of these properties are significantly lower than those conventionally expected for rocks. Melosh further suggested that acoustic fluidization is the process that permits effective strengths to be so small. In this model, the intrinsic rock or debris rheology is described by a standard friction law, but the presence of strong sound waves generated by the impact causes stochastic decreases in confining pressure, allowing sliding.

These results point to a mechanism that may qualitatively explain why large basins on Venus and the Earth lack mantle uplifts, unlike at least son e on the Moon and possibly Mars. As with many other aspects of crater morphology (most notably the diameter of the simple-complex transition), we suggest that surface gravity controls the magnitude of this deep structural uplift as well. It is well-known that the intrinsic frictional strength of a rock mass is proportional to overburden pressure (*e.g.*, Brace and Kohlstedt 1980). Therefore acoustic fluidization is most effective near the surface where sound pressures are arge and overburden pressure is sma l. This allows crater collapse to proceed nearly independently of planetary size Iowever, decreasing acoustic pressure and increasing resistance to frictiona sliding limit ıe depth to which this process is effective 'or similar-sized craters (and hence magnitudes of the acoustic field), this depth will vary inversely with gravity (though perhaps not in direct proportion). Therefore we suggest that the depth of acoustic fluidization may have penetrated to well within the nantle for large basins on the Moon, reducing the effective strength so tha isostatic equilibrium was quickly attained. In contrast, such changes iı strength may have been confined to the crust for Venus and ıe Earth, so that gravitational driving stresses could not promptly overcome the frictional strength of even heavily fractured rocks near the crust-mantle boundary. Because this model predicts a continuous variation in the depth of strength reduction as a function of gravity, mantle uplifts on Mars should be intermediate. Because frictional strength is independent of rock type, target properties should not affect mantle uplift in this model.

Alternatively, target properties could play a role if differences between the creep strength of the crust and upper mantle are sensed during crater collapse. Due to the strongly nonlinear creep behavior of rocks, the high stresses and temperatures present at the onset of crater modification may lead to effective viscosities sufficiently small so that cavity collapse is partially hydrodynamic. Stress differences in particular are much larger than those normally encountered in geological processes (Melosh 1989, p. 176). In this case, the thickness of the crust may control mantle uplift. Although mantle closer to the transient crater would experience larger stresses tending to drive uplift, such movement is inhibited by the exponential temperature dependence of viscosity: a temperature decrease of several hundred degrees or more over a 40-70 km difference in crustal thickness may provide an increase in effective viscosity of many orders of magnitude. Therefore the upper mantle at ~80 km depth in the Moon could have been weak enough to be entrained in the cavity collapse flow, whereas the upper mantle at 10-40 km depth on Earth and Venus may have presented an effective barrier to flow. The thin-crust argument requires that temperatures still increase with depth immediately after an impact, or specifically that temperatures at the base of the lunar crust are greater than those at the base of the terrestrial or Venusian crusts. This condition is likely to be met as it is not unreasonable to assume that thermal gradients on the Moon at 3.8 Ga were comparable to those on Venus and Earth in the recent past, and the skin depth for heat deposition in even the large lunar basins Orientale and Imbrium are 20 km or less (Bratt *et al.* 1985b).

Finally, we note that the lack of mantle uplifts in smaller basins and craters may be explained as a simple effect of flow penetration depth, independent of absolute viscosity. The characteristic depth of flow in an incompressible fluid half-space driven by surface perturbations with wavelength λ is approximately given by the skin depth $d = \lambda/2\pi$. Such terms appear regardless of Reynolds number, *i.e.*, whether the flow is slow viscous relaxation or rapid hydrodynamic rebound. Thus for a characteristic wavelength $\lambda \sim 2D$ (where D is the crater diameter), d is roughly D/π. In order for isostatic rebound to occur, flow must penetrate to depths of order the crustal thickness. For crustal thicknesses on all the terrestrial planets of order tens of kilometers, impact structures must be of order 100 km in size and larger in order for cavity collapse flow to sense the mantle at all. A similar conclusion about the long-term isostatic response can be reached by considering the characteristic flexural wavelength of the lithosphere (*e.g.*, Turcotte and Schubert 1982).

CONCLUSIONS

Using a combination of spherical harmonic and line-of-sight analyses we have determined that Mead crater is essentially uncompensated, in contrast to most of the shorter-wavelength features on Venus, which appear to be compensated at a relatively shallow depth. Thus appreciable mantle uplift cannot have occurred either immediately after crater formation, or gradually over the subsequent period of time. The lack of immediate uplift requires that the yield strength in the upper mantle (under conditions appropriate to the collapse of the transient cavity) be quite high. This is in apparent contrast to the Moon, where mantle uplifts have been inferred beneath the mare basins. Longer term uplift must have been resisted by a thick mechanical lithosphere, which must have been at least 30 km thick, and perhaps much thicker.

We have also demonstrated that the present generation of high-resolution spherical harmonic gravity models can be successfully used to study small-scale features on the planets that previously required the exclusive use of LOS techniques.

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FIGURES

<u>Figure 1a.</u> Radar image of Mead crater and surrounding area. Mead is at the center of t e image, and the northern edge of A hrodite can be seen at the bottom. This figure was taken primarily from the Magellan C1–MIDR (Compressed-Once Mosaic Image Data Record) 15N060;1, with some edges and gaps filled using C1-MIDR's 00N043;201, 00N060;1, 00N060;202, 15N043;201, and 15N060;201, all of which were acquired in a left–looking configuration. The image is about 1600 km across. Note that all following contour map figures, with the exception of Fig. 9, cover the same area as shown in this figure

<u>Figure 1</u> opography of the same area as Fig. 1a. Shading differences represent 250 m contours, with darker areas lower and ighter areas igher. The deepest portion of Mead's floor is at a radius of 6050.5 km. ata was taken from the Magellan GTDR.3;1

<u>Figure 2.</u> Gravity profile computed at the surface for a simple topographic model f Mea crater istance is measured ra ially fr the center of e axisymmetric depression. Only harmonic terms through degree and order 60 ave been retained in h. gra i y c m utation

<u>Figure 3a.</u> Vertical acceleration map at spacecraft altitude (182 km) from the 60^{th} degree and order gravity field MGNP60SAAP (Konopliv and Sjogren 1994). In this and all following contour plots, solid and dashed contours indica e positive and negative values, respectively; the zero contour is denoted by a dot-dash line. The contour interval is 0.5 mgal. The size and

location of Mead crater is shown by the sol d circle near the center of the figure.

Figure 3b. Vertical acceleration map as in Fig. 3a, but computed at the surface.

Figure 4. Magellan LOS acceleration profiles over Mead crater derived from Doppler residuals with respect to a 21^{st} degree and order gravity model (McNamee *et al.* 1993), which should be able to resolve wavelengths down to 17 degrees of arc (about 1800 km). We have introduced an artificial 5 mgal vertical separation between orbits so that the individual profiles can be more easily seen. Mead crater is about 3° across, centered at 12.5° latitude. A narrow dip in acceleration associated with Mead can be seen in all profiles except 6197.

Figure 5. Magellan LOS acceleration profile over Mead crater derived from Doppler residuals of orbit 6186 with respect to GM only (corresponding to a spherically symmetric reference field) and with respect to a 36^{th} degree and order gravity model (Nerem *et al.* 1993) with a wavelength resolution of 10 degrees of arc (~1000 km). Whereas the background signal differs considerably between these wo reducti ons, the anomaly over Mead remains almost unchanged.

Figure 6. Mage lan LOS acceleration profile ov 'er Mead crater derived from Doppler residuals of orbit 6186 with respect to a 60^{th} degree and order gravity model (Konopliv and Sjogren 1994) with a wavelength resolution of 6 degrees of arc (~600 km). With this reduction the background signal is nearly

eliminated (< 0.2 regal) and the strength of the Mead anomaly decreases to about 1.0 regal (1.6 mgal peak- to–peak).

Figure 7. Contour map of LOS Doppler acceleration residuals with respect to the 60^{th} degree and order gravity model, derived from of all orbits with good X-band data between orbit numbers 6171 and 6233. Contour interval is 0.5 mgal. Vertical lines show the location of the orbits used and the circle at the center of the map denotes the size and location of Mead.

Figure 8. Topography from the 360^{th} degree fit to the global altimetry data. Contour interval is 200 m.

Figure 9. Gravity field at spacecraft altitude to degree 60, computed using only the local topography from a 5° square centered on Mead. The size and location of the crater can be seen at the center of the figure. Contour interval is 0.25 regal. Note the smaller area (5°× .5" versus 15°× 15°) of this map relative to that of the previous figures.

Figure 10. Generalized isostatic anomaly map of the Mead region. Contour interval is 0.5 regal.

Figure 11. Spectral admittance for Venus (from Konopliv and Sjogren 1994). Also shown are the various apparent depths of compensation implied by the admittance at various wavelengths.

Figure 12. Isostatic anomaly map, computed at the surface for a compensation depth of 25 km. Mead crater shows as a -27 regal anomaly at the center of the figure. Contour interval is 2 mgal.

Figure 13. Calculated line of sight accelerations for r it 6186 from the filtered opogra hy vithin ? crater radii t Mead (see ext), assuming no compensation Peak eak amplitude and ar abou 5.5 mgal and 5 degrees f arc, r s ectively

Figure 14. The pper curve shows the calcu ed vertical displacement of the lithosphere as a nc ion of osphere thickness beneath a 270 km diameter cylindrical crater vi h an in ial depth r 000 m The lower c rve shows the ratio of this displacement the inal r depth ini ial dep minus displacement)